doi:10.1038/nature16147

Potential sea level rise from Antarctic ice sheet instability constrained by observations: Supplementary Information

Contents

1 Supplementary Methods

1.1 Ice sheet model

We use GRISLI (GRenoble Ice Shelves and Land Ice model)[1,2], a 3D finite difference thermomechanically coupled ice sheet model. In the grounded ice domain, it solves the mechanical equations using the shallow ice approximation for vertical shearing and the shallow shelf approximation as a sliding law and for ice shelf flow. This type of model, often called hybrid shallow ice-shallow shelf, includes the minimum physics to account for longitudinal stress transmission, which plays a crucial role in the inland propagation of dynamical coastal changes. We use a 15 km horizontal resolution.

1.1.1 Enhancement factor

The enhancement factor E_f represents the impact of ice anisotropy and was first introduced to calibrate simulated ice sheet volume[3,4]. It appears in the Glen flow law relating strain rate to stress deviator:

$$\dot{\epsilon}_{ij} = E_f \; A_T \; \tau^{n-1} \; \tau_{ij} \; ,$$

where $\dot{\epsilon}_{ij}$ is a strain rate component, τ_{ij} the corresponding deviatoric stress component, τ the effective shear stress and $n = 3$. The temperature-dependent rheological coefficient A_T is derived from laboratory measurements[5]. Separate enhancement factors are used for the shallow ice and shallow shelf approximations (Section 1.5): the typical fabric (vertical C-axes) found in ice sheets tends to enhance shearing in the vertical plane, i.e. the part of ice flow described by the shallow ice approximation ($E_f^{SIA} > 1$), but tends to make ice stiffer under the shallow shelf stress regime ($E_f^{SSA} < 1$).

1.1.2 Basal friction law

Basal friction in the model represents the combined effects of subglacial hydrology and rheological properties of the underlying sediment, and significantly affects the sliding law. This sliding law, or basal friction law, relates basal velocity ($\mathbf{U}_b = (U_{b,x}, U_{b,y})$) and basal friction ($\tau_{\mathbf{b}} = (\tau_{b,x}, \tau_{b,y})$) at horizontal location $\mathbf{x} = (x, y)$ and time t:

$$\tau_{b,x}(\mathbf{x}; t) = -\beta'_m(\mathbf{x}; t) \; U_{b,x}(\mathbf{x}; t)$$

and similarly for $\tau_{b,y}(\mathbf{x}; t)$, in which $\beta'_m(\mathbf{x}; t)$ is an *effective basal drag coefficient* (Pa a m^{-1}):

$$\beta'_m(\mathbf{x}; t) = \beta(\mathbf{x}) \; \frac{|\mathbf{U}_b(\mathbf{x}; t)|^{\frac{1}{m}-1}}{|\mathbf{U}_b(\mathbf{x}; t = t_0)|^{\frac{1}{m}-1}}$$

where $\beta(\mathbf{x})$ is the basal friction coefficient in the linear viscous formulation of the sliding law (see below). The shallow ice and shallow shelf approximations are combined wherever $\beta'_m(\mathbf{x};t)$ is lower than 5×10^6 Pa a m^{-1}; for larger values, sliding is not significant and the velocity is set to zero.

We use three alternative values for the exponent m corresponding to three basal friction laws (Section 1.6): $m = 1$ for a linear viscous law, where basal friction is proportional to basal velocity; $m = 3$ for a Weertman nonlinear power law; and $m = 100$ to approximate a plastic law, where the magnitude of basal friction is independent of basal velocity.

The most appropriate basal friction law depends on the local characteristics of the bed, which varies from soft sediment to hard bedrock but is generally poorly known. For example, modelling of Pine Island and Thwaites Glaciers constrained by observed velocities indicates the bed in this region varies from soft sediment to hard bedrock and back on length scales of tens of kilometres to kilometres[6]. Areas of soft sediment may be governed by either linear-viscous deformation or plastic failure: early seismic observations of sediment deformation supported a linear law[7,8], and ice sheet models often use this form[9] (in part because it is the most straightforward to implement: for example to infer $\beta(\mathbf{x})$, Section 1.5), while later laboratory measurements of basal till[10,11] support plastic flow. Areas of hard bedrock, which may be in the form of rough 'sticky spots', can be described with a Weertman law. Modelling of Pine Island and Thwaites Glaciers supports a nonlinear law in regions of hard bedrock and a plastic law for deforming sediment[6], and finds that a linear-viscous law underestimates sensitivity to changes in basal friction near the grounding line.

For the linear friction law ($m = 1$), the effective basal friction coefficient is constant in time, $\beta'_1(\mathbf{x};t) = \beta(\mathbf{x})$. We derive this spatial field with an inverse method (Fig. ED2a; Section 1.5). There are not enough time-dependent velocity data at the scale of the whole ice sheet to infer an initial map $\beta'_m(\mathbf{x};t = t_0)$ for $m > 1$ so we use β'_1 for the first time step (then evolve the field according to the equation above).

The coefficient $\beta'_m(\mathbf{x};t)$ is used in the new grounding line retreat parameterisation described in the next section.

1.2 Retreat parameterisation

Grounding line position is a sub-grid characteristic, not a state variable of the model. We implement a new parameterisation of grounding line retreat rate $v(\mathbf{x};t)$ (km a^{-1}) as a piecewise linear function of the logarithm of the effective basal friction coefficient, $\alpha(\mathbf{x};t) = \log_{10} \beta'_m(\mathbf{x};t)$, as follows:

$$v(\mathbf{x};t) = \begin{cases} v_{max}, & \text{for } \alpha(\mathbf{x};\mathrm{t}) \leq \alpha_{\mathrm{low}}, \\ v_{max} + (\alpha(\mathbf{x};t) - \alpha_{low})\ \frac{(v_{max}-v_{min})}{(\alpha_{max}-\alpha_{min})}, & \text{for } \alpha_{\mathrm{low}} < \alpha(\mathbf{x};\mathrm{t}) < \alpha_{\mathrm{high}}\ , \\ v_{min}, & \text{for } \alpha(\mathbf{x};\mathrm{t}) \geq \alpha_{\mathrm{high}}. \end{cases}$$

where $\alpha_{low} < \alpha_{high}$ and $v_{min} < v_{max}$, so lower values of $\beta'_m(\mathbf{x};t)$ result in faster retreat. Grounding line retreat in each sector is initiated from the year of onset for that sector (Fig. ED1a; Section 1.6.2) and continues in every subsequent year (subject to glaciological conditions described in the next section). Retreat is implemented by prescribing a decrease in ice thickness for grounded grid cells contiguous to floating ice (Section 1.4).

Parameterising v in terms of the *effective* basal friction coefficient introduces a positive feedback for the Weertman and plastic laws ($m > 1$): grounding line retreat leads to higher velocities upstream, by reducing basal friction and increasing the surface slope; increased velocities lead to a decrease in β'_m, which increases the retreat rate (up to v_{max}). Figure ED2e shows one example: a map of α at 2200 in the ASE for the plastic sliding law ensemble member most successful in reproducing observations (maximum likelihood, Section 1.7). In the parts of Pine Island and Thwaites Glaciers that are not ungrounded by 2200, β has reduced by an order of magnitude (α decreased by 1).

More details of the parameterisation implementation and interpretation are given in Section 1.4.

1.3 Glaciological conditions for retreat

1.3.1 Schoof flux condition

We permit grounding line retreat only over bedrock that is below sea level and down-sloping inland, which has long been considered topographically unstable. The only exception is to allow retreat over small bumps using a 'Schoof flux' condition that depends on basal friction.

Retreat can occur if the ice flux across the grounding line computed by the model is less than the analytical flux q_g proposed by Schoof[12] assuming no buttressing:

$$q_g = \left[\frac{\bar{A}(\rho_i g)^{n+1}(1-\rho_i/\rho_w)^n}{4^n \beta} \,, \right]^{\frac{1}{1/m+1}} H_g^{\frac{1/m+n+3}{1/m+1}} \,,$$

where ρ_i and ρ_w are the densities of ice and water, g is acceleration due to gravity, H_g is ice thickness at the grounding line, and depth-averaged rheological coefficient (for $n = 3$) is:

$$\bar{A} = \left[\frac{H}{\int_H (E_f^{SSA}\ A_T)^{-\frac{1}{3}}\ dz} \right]^3 .$$

Note that Schoof uses a sliding law exponent definition m that is equivalent to our $1/m$.

This condition allows grounding line retreat to occur in more locations for the Weertman and plastic basal friction laws ($m > 1$) than for the viscous ($m = 1$).

1.3.2 No suction check

We also ensure thinning due to grounding line retreat does not exceed the maximum permissible rate, using theoretical knowledge of maximum possible stresses at the grounding line that we call the 'no suction check'. It is strongly dependent on ice thickness and is essentially active in the regions where bedrock is not far below sea level (George V Land, for instance).

Ideally we would calculate back-stress from ice shelves explicitly, but this is a particular challenge in modelling grounding line migration because it requires simulation of the ice shelf-ocean-atmosphere system. Instead we ensure tensile stresses do not exceed those from buttressing by water alone, which we call the free-water tensile stress, and calculate the maximum corresponding strain-rate, expressed as a maximum thinning rate.

Taking without loss of generality $U > 0$ and $\partial H/\partial x < 0$, the free-water tensile stress in one dimension reads

$$\tau_{xx}^{F} = \frac{\gamma}{4} H,$$

where $\gamma = (1 - \rho_i/\rho_w)\,\rho_i g$. The free-water tensile strain-rate follows,

$$H\,\frac{\partial U}{\partial x} = \bar{A}\left(\frac{\gamma}{4}\right)^{n} H^{n+1}\ .$$

Using the mass conservation equation, the condition on maximum strain rate is

$$\frac{\partial H}{\partial t} = \mathrm{smb} - \frac{\partial \mathrm{UH}}{\partial \mathrm{x}},$$

which is expressed as a maximum thinning rate. For two horizontal dimensions, the maximum thinning is the sum of those for x and y.

1.4 More on the retreat parameterisation

This section describes the motivation for the parameterisation functional form, its implementation with the glaciological constraints, and its physical interpretation.

1.4.1 Choice of functional form

We parameterise retreat rate, rather than fluxes or elevation changes, because this is a one-dimensional observable quantity. The dependence of retreat rate on effective basal friction coefficient cannot be empirically derived, because the latter can only be inferred from surface velocity observations (Section 1.5). Our parameterisation is therefore designed to be qualitatively consistent with the theoretical analysis of Schoof[12]. The Schoof flux (SF) has this type of dependency, where ice flux through the grounding line increases if the bed is more slippery. We do not use the SF to compute retreat rates directly, because it is a steady state analysis and does not take spatial variation of β into account; instead we use it as an analytical limit, where retreat is allowed if the flux computed by the model is less than the SF.

Our piecewise linear dependency of the parameterisation is motivated by the simplest functional form that is qualitatively consistent with the SF:

1. As β decreases, the SF increases. The simplest parameterisation is therefore a linear inverse dependency, and we choose this for the central part of the function.

2. At $\beta = 0$, the SF is infinite. This is clearly unrealistic so a limit is needed. We choose the simplest possible, a threshold v_{max}.

3. At high β, there is no sliding (Section 1.1.2). Here the SF is only due to shearing (shallow ice approximation) and is very similar to the model flux so the grounding line is generally unlikely to move. We choose $v_{min} = 0.1$ km a^{-1}, rather than zero, to avoid missing possible retreat: the grounding line can sometimes advance past a region of high β, particularly with the nonlinear and plastic sliding laws, if velocity increases propagate upstream, decrease β and initiate sliding.

Higher order functions than piecewise linear would be difficult to justify, given current understanding. But the perturbed parameter sampling of the piecewise linear form (Section 1.6) allows us to scan a wide range of possibilities with a behaviour that is consistent to first order with the Schoof analysis.

1.4.2 Interaction with glaciological conditions

Retreat is implemented by applying the following procedure at each time step:

- Along every segment that includes the grounding line (one grid cell floating and one grounded) the sub-grid grounding line position is diagnosed from ice thickness and bedrock elevation (assuming both vary linearly). Diagonal segments are also tested because simulations with a finer grid (5 km) indicated this made the procedure less dependent on grid size and avoided rectangular patterns in the retreat.

- If the bedrock is downsloping toward the grounded grid cell or the Schoof flux condition is passed, the prescribed retreat rate is translated into a rate of decrease of the height above flotation in the grounded grid cell. For a given grounded grid cell, retreat could come from multiple directions so we use the largest thinning rate obtained.

- The no suction check is applied, and finally the new ice thickness is computed. This value is prescribed as a Dirichlet condition in the mass conservation equation.

- If flotation is achieved during this procedure, the grid cell is marked as ice shelf, with thickness fixed at flotation for the rest of the simulation.

The result of the glaciological conditions (Schoof flux condition and no suction check) is a hybrid, statistical-physical approach in which the simple parameterisation of retreat onset and rate is modified using process modelling of ice dynamics under the basal boundary conditions. In particular, although the maximum retreat rate parameter v_{max} is a constant (for a given ensemble member: Section 1.6), and retreat is 'requested' continuously after the onset date, the actual retreat may be slower, stop, or show episodic behaviour, because it is modified by the implications of current local conditions for physically plausible ice flux and extension rates at the grounding line.

1.4.3 Interpretation

This parameterisation replaces the explicit simulation of changes in buttressing, whether from ice shelf collapse or basal melting, by representing the effect of reduced backstress on the grounding line. Retreat onset effectively corresponds to buttressing being sufficiently reduced to trigger retreat, while the retreat rate parameterisation describes the dependence of grounding line migration on basal conditions. We parameterise retreat and stabilisation, not advance, because the projected drivers (ice shelf losses and new intrusions of Circumpolar Deep Water) act in this direction for the current climate-ice sheet system and because we wish to assess potential sea level rise in the event of MISI. The parameterisation is calibrated with observations of ASE ice losses (Section 1.7).

As changes in ice shelf buttressing are not explicitly simulated, initial ice shelves are retained and (as described above) newly formed shelves are prescribed at the maximum ice thickness that allows flotation, thus providing an upper bound for ice shelf buttressing. Test simulations in which ice shelf collapse is simulated (by reducing the viscosity until the backstress is very small) showed no substantial differences, because the parameterisation and glaciological conditions control the retreat more strongly. For example, if the prescribed retreat rate is low, removing the ice shelves does give a higher initial model ice flux, but the Schoof flux condition is less likely to be met at bedrock bumps so retreat slows or stops until the flux is sufficiently small. If the prescribed retreat rate is high, retreat is restricted in any case by the no suction check.

Similarly there is also no calving law. Our method to prescribe retreat rates has some similarities with the recently proposed hydrofracturing mechanism [13] and has the same order of magnitude of rates. The main differences are that the fracturing parameterisation is expressed in terms of a 'horizontal wastage rate', that it is only applied to cliffs (surface elevation 100 m above sea level, typically in regions of deep bedrock), and that we keep ice shelf fronts at the present positions. Our parameterisation also acts in areas of shallow bedrock, though retreat is limited in these regions by the no suction check so this difference may not be very large.

1.5 Boundary and initial conditions

The boundary conditions are maps of bedrock elevation, surface mass balance, and geothermal heat flux[14,15,16], averaged from 5 km resolution onto the 15 km grid. The initial state of the ice sheet is determined with an initialisation procedure that includes adjustment of the enhancement factors and the basal friction coefficient field to improve the match with observations and minimise spurious changes.

Initialisation is a compromise between several criteria: surface elevation (S) and surface velocities must be close to observations; surface elevation change ($\partial S/\partial t$) must be of the same order of magnitude as altimetry observations (our target was $\partial S/\partial t = 0$); and velocity and temperature fields must be compatible (have been coupled for a few centuries).

The procedure requires several steps. We first run the heat equation under present climatic conditions until a steady state is obtained, with an initial guess for the velocity field and the geometry initialised at present day observations then allowed to relax for 10 years. This means we do not explicitly take into account thermal memory due to the glacial-interglacial transition, but this effect is partly included in the enhancement factor calibration, and the trend it could produce is insignificant over our two century projection.

In the second step, an initial map of $\beta'(\mathbf{x})$ is derived from inversion of observed surface velocities with the ice sheet model Elmer/Ice using a linear basal friction law[17]. This is prescribed as a boundary condition in GRISLI and the model run for 200 years starting from present day geometry. After 10 years of simulation, the velocity field is already very similar to the observations but $\partial S/\partial t$ values are large, in some places up to 10 m a^{-1}. This is mainly due to the resolution of GRISLI being much coarser than that of Elmer/Ice. In some regions it is also due to a lack of bedrock elevation observations. At the end of this simulation $\partial S/\partial t$ is acceptably small, but the ice sheet geometry is rather different from the observed value at some locations. In practice, ice fluxes at the end of this simulation are close to balance fluxes on the GRISLI grid (including the staggered features of the grid), so we use them as a new target to derive a second $\beta'(\mathbf{x})$ map more compatible with GRISLI.

Finally another simulation starting from present day geometry is performed to improve relaxation. This procedure is repeated for various values of shallow ice enhancement factor and we choose the set $\{E_f^{SIA}, \beta'(\mathbf{x})\}$ that minimises the trend in volume above flotation for each of the 14 drainage basins and the whole of Antarctica ($E_f^{SIA} = 1.8$). E_f^{SSA} is calibrated to obtain realistic velocities at the front of the Ronne-Filchner and Ross ice shelves: the resulting ratio of enhancement factors is $E_f^{SIA}/E_f^{SSA} = 4.8$.

After 200 years the selected simulation fulfils our requirements for a reasonable initial state. The mean difference between simulated and observed surface elevations (Fig. ED2a) is 22 cm and the standard deviation 45 m; Figure ED2b shows the initial simulated velocities and Figure ED2c the initial map of $\beta'(\mathbf{x})$. All simulations are spun-up for a further five years from this initial state, to adjust to the specified bedrock topography map and basal friction law (Section 1.6).

To quantify remaining model drift, we perform a reference simulation for each basal friction law with the grounding line fixed at the present day position. The change in surface elevation from 2000 to 2200 has a mean of 8.6 cm and a standard deviation of 12.1 m for the viscous law reference simulation, -6.1 cm and 11.8 m for the Weertman, and -60 cm and 11.2 m for the plastic. All projections are expressed as the difference between a given experiment and the corresponding reference run.

1.6 Ensemble design

We generate 3000 model versions by varying 16 uncertain model inputs: (i) exponent of the friction law m; (ii) retreat rate parameters $\{v_{max}, \alpha_{low}, \alpha_{high}\}$; (iii) retreat onset dates for each sector $\{t_1, ..., t_{11}\}$; and (iv) bedrock elevation map. We construct this by generating 1000 parameter sets for inputs (ii-iv) and repeating these for each of the three values of m. We simulate two hundred years (2000-2200) with each model version.

1.6.1 Retreat rate parameters

We fix v_{min} in the retreat parameterisation at 0.1 km a^{-1} and sample the remaining uncertain parameters from independent uniform distributions:

$$\begin{aligned} v_{max} &\sim \mathrm{U}(0.2, 3) \text{ km a}^{-1} \\ \alpha_{low} &\sim \mathrm{U}(0, 5) \\ \alpha_{high} &\sim \mathrm{U}(3, 7) \end{aligned}$$

using a 'maximin' Latin Hypercube (space-filling design that maximises the smallest distance between any two points) with the constraint that $\alpha_{low} < \alpha_{high}$. The resulting piecewise linear functions are shown in Fig. ED1b. The maximum value of v_{max} is based on observed[18] surges of the Pine Island Glacier grounding line of up to 2.8 km a^{-1}. Higher values of retreat rate are restricted by the no suction check (Section 2.2.1). The maximum value of α_{high} approximately corresponds to the value at which basal sliding no longer occurs (Section 1.1.2).

1.6.2 Retreat onset distributions

We must parameterise the triggers of MISI in this scheme, because ocean-ice feedbacks and buttressing are not simulated explicitly. We assign probability distributions of retreat onset dates for each of 11 sectors (comprising the 14 drainage basins), which we take to be independent. Our projections are therefore conditional on the basin scenarios, which translate the best available expert judgement into probability distributions. This is a novel approach compared with recent studies of MISI that parameterise basal melt rates and margin weakening[19,20], because it allows assessment of risk similar to that for natural hazard events, where risk of MISI = probability × impact.

Our use of retreat onset distributions is similar in philosophy to the discharge growth rates and constant collapse probabilities assumed by Little et al.[21] from a combination of observations, model projections and expert elicited scenarios, though in our study the assigned probabilities are not constant in time, 'retreat onset' only leads to ice losses if it is found to be dynamically possible in the model, and losses are also modified by ice dynamical constraints.

The probability distributions (Fig. ED1a) are assigned by expert assessment of observed grounding line retreat and thinning[22,23,18], and regional model projections of ocean temperatures, ice shelf basal mass losses and firn air depletion driven by the HadCM3 general circulation model under A1B[24,25,26,27], by setting a probability for each twenty year time period of each basin. The experts consulted by the authors were David Vaughan, Hartmut Hellmer, Peter Kuipers Munneke, Michiel van den Broeke and Hamish Pritchard. Broadly, the assigned cumulative probability of onset reaches one by 2030 for basins where MISI is thought to be underway or very likely, approaches one by 2200 for basins where surface melt ponding on ice shelves is projected during this time (under the assumption that melt ponding will lead to shelf collapse and that is sufficient to trigger MISI), and is lower for basins with only projected increased ocean melting or other evidence of vulnerability (as this may not be sufficient to trigger MISI). More details are given below, and sensitivity of the projections to these distributions is discussed in Section 2.2.2.

Peninsula, Amundsen Sea and Marie Byrd Land. Grounding line retreat or dynamic thinning has already been observed[22,23,18]. These basins are assigned cumulative probabilities that reach one by 2030 (i.e. retreat is triggered in all 3000 simulations by this date).

Dronning Maud, MacRobertson and Princess Elizabeth. Some thinning has been observed[23], and firn air depletion has also been projected over the ice shelves by a firn model driven by the RACMO2 regional atmospheric model under A1B[25], particularly in the second century. The cumulative probabilities assigned for each basin are assigned to approach one by 2200, with an increase during the 22^{nd} century. The assessments also incorporate projections of ocean temperatures and ice shelf basal mass loss, which have been predicted to increase by the high resolution sea ice-ocean model FESOM (Finite-element Sea ice-Ocean Model) and the lower resolution BRIOS (Bremerhaven Regional Ice-Ocean Simulations) when both models are driven by HadCM3 under A1B[27]. For Princess Elizabeth Land, both models project increased ocean temperatures during the 21^{st} century, followed by more rapid warming in the 22^{nd} (same simulations: Timmermann and Hellmer, pers. comm.). This basin is assigned increasing probability in the 21^{st} century then a more rapid increase from the start of the 22^{nd}. For MacRobertson Land ("Amery" in the reference), FESOM projects increased ice shelf basal mass loss during the 21^{st} century followed by more rapid losses in the

22^{nd}, while BRIOS projects a temporary increase in basal loss only during the 21^{st} century. This basin is assigned increasing probability of retreat onset in the 21^{st} century with a faster increase from the middle of the 22^{nd}. For Dronning Maud Land ("Eastern Weddell Ice Shelves: EWIS"), FESOM projects increased ice shelf basal mass loss during the 21^{st} century followed by more rapid losses in the 22^{nd} and further acceleration at the end of that century, while BRIOS shows little increase. This basin is assigned increasing cumulative probability through the two centuries with more rapid increase at the end of the century.

Ronne-Filchner sector (Ellsworth, Pensacola Mountains and Shackleton Range). FESOM and BRIOS forced by HadCM3 under A1B[24,27] ('FRIS' in the second reference) project increased ocean temperatures and ice shelf basal melting, particularly from the late 21^{st} century (BRIOS) and 2100 (FESOM), due to new intrusion of Circumpolar Deep Water in the region. These basins are assigned common retreat onset dates due to their common forcing, with more rapidly increasing probability from around 2100.

Wilkes Land and George V Land. Thinning has been observed[23] and George V Land is also thought vulnerable to MISI[26]. These are assigned increasing probabilities throughout the two centuries that do not approach one by the end, with higher probabilities assigned to Wilkes Land due to the observed dynamic changes in Totten Glacier.

Ross sector (Siple Coast and Transantarctic Mountains) and Enderby Land. FESOM projects increased ocean temperatures and ice shelf basal mass loss in the Siple region, while BRIOS does not[27]. For Enderby Land there are few significant ice streams, few ice shelves, and no other projected MISI triggers. These three basins are assigned distributions that approach 10% cumulative probability by the end of the second century, with identical onset dates for the two Ross basins.

1.6.3 Bedrock topography

The effect of uncertainty in bedrock elevation is assessed with a geostatistical approach by generating multiple perturbed bedrock maps. Measurement uncertainty in elevation is not well-quantified, so each sector is assigned to one of four categories, 'Very Good', 'Good', 'Poor', or 'Very Poor', based on the BEDMAP dataset[28] and later improvements[29,30]. The sectors are assigned as follows: 'Very Good' for Amundsen Sea, Siple Coast; 'Good' for the Peninsula, Ellsworth, Dronning Maud Land and Enderby Land; 'Poor' for the Pensacola Mountains, MacRobertson Land, George V Land and the Transantarctic Mountains; and 'Very Poor' for the rest. Measurement errors are assigned in the form of 2σ estimates motivated by BEDMAP: $2\sigma_{VG} = 50$ m, $2\sigma_G = 100$ m, $2\sigma_P = 250$ m, and $2\sigma_{VP} = 500$ m. We generate 100 isotropic Gaussian spatial fields with these standard deviations, using a squared exponential correlation function with error correlation length (distance at which error correlations fall to 5%) set to the correlation length of the elevations themselves, $l = 70$ km[28]. We add these perturbations to the observed

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bedrock map in regions of grounded ice while requiring bedrock elevation to be at least 100 m lower than the ice surface. Using maps with alternative correlation lengths (10, 40, 100 and 130 km) has little effect on the results. We sample randomly from the 100 bedrock elevation maps for the 1000 parameter sets.

1.7 Observational calibration

We use observed mass losses during 1992–2011 from the reconciled synthesis of satellite altimetry, interferometry, and gravimetry data sets[31], using estimates for the Amundsen Sea Embayment because it is the only region in which substantial grounding line retreat has been observed. For each sub-basin we calculate the mean rate of change in mass M from altimetry and input-output methods (weights $1/\sigma_i^2$). We sum the Pine Island (PIG) and Inter-stream ridge (INT) sub-basins together (corresponding to the GRISLI Pine Island Glacier region, P), and the Thwaites (THW) and Haynes, Pope, Smith and Kohler (HSK) glacier sub-basins together (GRISLI Thwaites Glacier region, T), to obtain:

$$\frac{dM_o^A}{dt} = -59.0 \pm 3.7 \text{ Gt a}^{-1} \ .$$

where o denotes observations and A the Amundsen Sea Embayment. We assume this trend is due only to dynamic processes, i.e. that the mean change in surface mass balance for this region is negligible, based on the agreement between estimates[32] of increased outflow and decreased mass balance during the period 1996–2006.

We simulate only the later part of the observational period, since the initial grounding line position is from 2003-2004 and the acceleration of Thwaites Glacier mass loss – i.e. indicating MISI may have initiated – began in around 2006[33,34].

However, since mass losses have accelerated during the observational period (1992-2011)[33,34], we must correct the simulated mass changes. We define scaling factors as the ratio of the mass trend from 1992-2011 to the mass trend from 2004–2011 (for PIG) or 2006–2011(for TG),

$$f_P = \left(\frac{dM_{o'}^P}{dt}\right)_{1992-2011} / \left(\frac{dM_{o'}^P}{dt}\right)_{2004-2011}$$

$$f_T = \left(\frac{dM_{o'}^T}{dt}\right)_{1992-2011} / \left(\frac{dM_{o'}^T}{dt}\right)_{2006-2011}$$

where o' denotes time-dependent estimates of mass change from radar altimeter data[33] from 1992-2013 (Fig. 4 in that study; GRISLI Thwaites basin is Thwaites plus PSK). This gives $f_P = 0.56$ and $f_T = 0.63$. The corrected trend for the ASE is then

$$\left(\frac{dM_m^A}{dt}\right)_{1992-2011} = f_P \left(\frac{dM_m^P}{dt}\right)_{y_1,\ldots,y_8} + f_T \left(\frac{dM_m^T}{dt}\right)_{y_1,\ldots,y_6}$$

where $y_1, ...y_n$ denotes the first n years after retreat onset in the basin.

The difference between each simulated ASE mass trend and the observation yields a likelihood value, calculated using a Gaussian likelihood function with mean dM_o^A/dt and variance $\sigma_t^2 = \sigma_o^2 + \sigma_m^2$ comprising observation error σ_o and GRISLI model error σ_m. Model error is an uncertain parameter (also referred to as structural uncertainty) that quantifies our tolerance of the limitations of GRISLI and avoids a 'perfect model' assumption. We set σ_m conservatively, by tolerating a 5% probability of the model simulating mass *gain* over this period and using Cantelli's inequality to calculate the corresponding σ_t (the width of the combined error distribution). This choice gives $\sigma_t = 13.5$ Gt a^{-1}, and thus $\sigma_m = 13.0$ Gt a^{-1} which is 22% of the signal.

This is conservative because the observations very clearly show mass loss (dM_o^A/dt is around -16 σ_o) and because it makes minimal assumptions about the shape of the error distribution (the Cantelli bound applies to any distribution with a finite second moment). Only 1% of the ensemble have mass losses smaller than $dM_o^A/dt + 3\sigma_t$ (to the left of the lefthand dot-dash line in Fig. ED5).

The Bayesian update is performed by applying normalised likelihoods as weights to each ensemble member. Likelihoods are normalised across each of the 1000 member parameter sets (Section 1.6), rather than all 3000, since we do not wish to calibrate the basal friction exponent m. In other words, we apply the calibration to each friction law sub-ensemble then average their posterior distributions. This is because the friction law varies spatially by bed type, and the ASE bed is not representative of the whole ice sheet, so we do not wish to give greatest weight to the friction law that best describes this region.

The observations therefore calibrate only the three parameters that control the retreat parameterisation (v_{max}, α_{low} and α_{high}, though the latter only weakly: Fig. ED8). They do not calibrate bedrock topography (Fig. ED8b, d), because this is a set of 2D fields with random noise added, rather than a systematically varied scalar parameter, nor retreat onset dates, because the calibration uses a fixed number of simulation years after PIG and Thwaites retreat (so the ASE retreat onset date is not used, and the other basins are not used at all). The sliding law exponent is also not calibrated, because calibration is performed for the individual sliding laws and then the three posterior distributions combined.

We estimate quantiles from the empirical cumulative distribution functions, and densities (for modes and figures) with kernel density estimation (fixed bandwidths 5 cm sea level equivalent for Antarctic sea level contribution; 0.05-2 cm and 0.1 mm a^{-1} for basin and glacier sea level contribution and rate; 0.3 km a^{-1} and 0.5 for joint density estimation of v_{max} and α_{low}).

2 Supplementary Discussion

Here we compare with other studies and test the sensitivity of individual simulations and 95% quantiles at 2200 to: (1) the Schoof flux and no suction check; (2)

prior distributions for retreat onset; (3) prior distributions for the retreat parameterisation; (4) choices in the observational calibration. We finish with a summary of these and some potential future directions.

2.1 Comparison with other studies

Our results for seven West Antarctic basins are consistent with regional projections to 2200 from the high resolution ice sheet model BISICLES[35] driven by the same ocean simulations on which we base our onset probabilities (FESOM and BRIOS driven by HadCM3 under A1B; Section 1.6.2). The most direct comparison, projections of dynamic ice loss forced only by FESOM (i.e. no accumulation changes), correspond to 19-30% quantiles at 2200 for the ASE (using three initial accumulation rates); 56-65% for Ellsworth, Pensacola Mountains and Shackleton; and 90% at 2200 for Siple Coast and Transantarctic Mountains (for which sensitivity tests are described below). Grounding line retreat lies within our 5% contour (Fig. 1). Comparing with the dynamic contributions for all FESOM and BRIOS A1B projections in that study (i.e. subtracting the accumulation, as the two are essentially independent; Section 2.3) broadens the range to 55-68% for the Ronne-Filchner basins and 78-90% for the Ross basins.

For the final basin, Marie Byrd Land, the FESOM-only projection is lower than our lowest ensemble member, but the contribution of this basin to our result is small (95% quantile 0.9 cm at 2200; all values sea level equivalent) and dynamic contributions from the other projections (as above) extend up to our 81% quantile.

Increasing Siple Coast and Transantarctic Mountain retreat onset probabilities ten-fold increases 95% quantiles at 2200 by 7.8 cm and 1.8 cm respectively. This lowers the BISICLES FESOM-only quantiles from 90% to 80% (Siple) and 83% (Transantarctic Mountains), indicating we have spanned a sufficiently large range with this test.

Our projections are essentially incompatible with upper bound estimates[36,37] of around 50-80 cm by 2100 and 140 cm by 2200 derived from physical arguments, simple extrapolation and low resolution numerical models and around 1 m by 2100 (95% quantile) from expert elicitation[38]. A statistically-based projection[21] assuming ASE collapse in 2012 and linear growth of ice discharge elsewhere has a higher 95% quantile (41 cm sea level equivalent from 1990 to 2099, including the negative contribution from increased surface mass balance), but we obtain a similar non-ASE 95% quantile (5-7 cm), and our ASE result lies within their range using alternative plausible discharge rates, so we consider our results consistent.

Rates of change in the ASE are also comparable with high resolution modelling studies. Projected rates from BISICLES under A1B[35] are generally consistent with ours, though their maximum rates occur later (22^{nd} century) than ours (mid-late 21^{st} century) (discussion in Section 2.3). Individual rates of change for Pine Island and Thwaites glaciers are consistent with multiple high resolution model results under idealised basal melting scenarios. For Pine Island, mass losses simulated after 50 years by three models (including BISICLES) under a high basal melting scenario[19] correspond approximately to our median, 66% and 95% quantiles (0.24, 0.33, and 0.65 mm a^{-1} at 2050). For Thwaites our estimates are generally higher than, but not inconsistent with, those from a basin model driven by various basal melting scenarios[20], which simulates rates in the lower third to half of our projections (up to 0.25 mm a^{-1} in the first century: corresponding to our 40% quantile at 2050 and 25% at 2100; up to 0.7 mm a^{-1} in the second century, just below our 33% quantile and median at 2150 and 2200 respectively). One Thwaites simulation in that study reaches 1 mm a^{-1} soon after the two centuries, similar to our 66% quantile at 2200 (0.97 mm a^{-1}). The wider range and tendency to higher values for Thwaites in our ensemble may arise from our exploration of modelling uncertainties and our larger basin area, which includes Smith and Kohler glaciers[22] (lower part of Fig. ED3b).

2.2 Sensitivity analyses

2.2.1 Glaciological conditions

Here we show results from test simulations to quantify the effects of the Schoof flux condition and no suction check. We use members of the plastic sliding law sub-ensemble, because they show larger projected sea level changes than the other sliding laws and are more sensitive (e.g. basal friction coefficient evolution, Schoof flux condition). We focus on 'Max likelihood', the ensemble member that best matches the ASE observations, and 'Ensemble max', the largest contribution at 2200, to represent "best/typical" and "maximum" results respectively. These and other results are shown in Fig. ED10.

If the Schoof flux condition is off, the grounding line cannot pass any increase in bedrock elevation. When switching this condition off in the simulations, the contribution at 2200 reduces from 47 to 27 cm (-42%) in the maximum likelihood member and from 152 to 36 cm (-76%) in the ensemble maximum. Maps of the surface elevation change in the ensemble maximum with and without the Schoof condition are shown in Figures ED4a and ED4b; the latter shows thinning but no grounding line retreat in Pine Island Glacier, and only limited thinning and retreat elsewhere.

When switching the 'no suction' check off the contribution at 2200 increases from 47 to 55 cm (+19%) for 'Max likelihood' and from 152 to 230 cm (+52%) for 'Ensemble max'. The violation of the theoretical limit on the rate of ice loss allows more connections to be made between retreated areas of the West Antarctic ice sheet on this timescale, more retreat in the Shackleton basin, and more retreat along the Wilkes ice stream in George V land which is considered vulnerable on long time scales[26] (Fig. ED4c).

Changes to sea level contributions for the nonlinear sliding law are 65-83% of those for the plastic law at 2100, and 55-91% at 2200. In general the nonlinear law gives relatively similar results to the plastic in terms of grounding line position (e.g. Fig. ED3) but the plastic contributes more to sea level due to greater inland thinning. The linear sliding law was not tested.

2.2.2 Retreat onset distributions

We have performed sensitivity testing of the retreat onset priors for each basin. We focus on the 95% quantile at 2200 because the contributions are largest and because the upper tail of the distribution is of interest.

We assume for these purposes that basin responses are sufficiently independent to reweight their individual contributions, based on inspection of the changes in each basin during the period before retreat is initiated, i.e. induced by connections to neighbouring basins. The largest induced changes are in the Ellsworth and Siple Coast basins: up to about 4 cm and 1 cm by 2200 respectively. Other basins show induced changes $\sim$O(0.01) cm or less over the two centuries, some of which are mass increases due to local thickening or migration of the ice divide.

For most basins – all but the Amundsen Sea, Peninsula and Marie Byrd Land – we take an Importance Sampling approach, reweighting the prior distributions by the ratio of the target (new) to proposal (original) probabilities. The "low" scenario has 20% lower probability in each decade, and the initial "high" scenario 20% higher. This is modified slightly for MacRobertson and Princess Elizabeth: increasing retreat onset probabilities by 20% means the cumulative probability would exceed one so the increases in the last decades are restricted slightly. The results show that reducing retreat onset probabilities by 20% generally reduces the 95% quantiles at 2200 by much less than 20%: -3 to -9% for all but three basins. The exceptions are Shackleton Range, Siple Coast, and Transantarctic Mountains, which respond by about -20% (i.e. linearly) and are discussed further below. Increasing onset probabilities by 20% has an even less sensitive response: 3-7% for all but two basins, with Siple Coast increasing by 20% and the Transantarctic Mountains by 15%. All changes are less than 0.8 cm in magnitude.

For the Amundsen Sea, Peninsula and Marie Byrd Land, there are no "no retreat" runs to assign greater weight to, so we perform an equivalent "low" scenario by setting a random sample of 20% (distributed equally across all decades) to zero sea level contribution. These results also show little sensitivity in the 95% quantile, -1.3 to -2.3%. There is no equivalent to this for the "high" +20% scenario, but we perform two additional importance sampling tests that redistribute probability to the latest or earliest decade:

- Low: 'Late retreat', 2020-2030
- High: 'Early retreat', 2000-2010

These show similar magnitude changes to the -20% scenario: at most 4% (1.8 cm). The changes to the ASE posterior distribution are shown in Fig. ED9a: the most different is the first test, in which 20% of the simulations are set to zero ("Probability x0.8").

For 11 of the 14 basins, then, 95% quantiles at 2200 are quite robust to changes in retreat onset probability. (Some lower quantiles do show larger changes, due to changes in the posterior distribution shape: seen for example in Fig. ED9a). Other modelling uncertainties, such as sliding law and maximum retreat rate, have greater influence. In these basins the main impact on sea level occurs after retreat onset – after this basal conditions and glaciological constraints slow or stop retreat – so changes in onset time have a relatively small effect.

For the remaining three basins, which are mostly linear over ±20% changes in probability – Shackleton, Siple Coast, and Transantarctic Mountains – we test larger changes. With a doubling of retreat onset probability in all decades, the percentage increases are less than 100%: 25% for Shackleton (1.5 cm), 92% for Siple Coast (2.6 cm), and 77% for Transantarctic Mountains (0.6 cm). With a ten-fold (i.e. 900%) increase in retreat onset probability the responses are 273% for Siple Coast (7.7 cm) and 246% for Transantarctic Mountains: (1.8 cm). The second test is not possible for Shackleton as the cumulative probability would exceed one. The results for Siple and Transantarctic Mountains are approximations, as only around 300 of 3000 simulations retreat and are therefore available for importance sampling, but they indicate an end to linearity with increasing retreat onset probability. We note that the active part of the Transantarctic Mountains region could be seen as belonging to Siple Coast (Fig. 1). Changes to the posterior distributions for these three basins are shown in Figure ED9b-d.

Responses of these three basins under the above tests are therefore less than 3% of the continental total, with the exception of Siple Coast (11%). But the latter is only obtained by increasing the probability of retreat onset ten-fold. So the continental projections are not sensitive, unless new evidence were to indicate there is a substantial probability of MISI being triggered in this region by changes to, or under, the Ross ice shelf.

We have also run simulations with 'immediate onset' in all basins, which are described in the next section.

2.2.3 Retreat rate parameters

We have tested sensitivity to retreat rate parameters using a very extreme scenario we call "immediate onset": retreat onset in 2000 for the ASE, Peninsula and Marie Byrd Land basins, and in 2020 for all other basins, with a plastic sliding law. We set α_{low}, α_{high} and v_{max} to their maximum values in the experimental design (3, 7 and 3 km a^{-1} respectively), 'Max params' in Fig. ED10.

We first test the effect of increasing α_{low} to 6.9. This has no effect on the contribution at 2100 or 2200, indicating that our choice for the prior maximum is sufficiently large.

We then test the effect of also increasing v_{max} to 5 km a^{-1}, 'Extreme params' in Fig. ED10. The no suction check strongly limits the contribution at 2200 when increasing v_{max} beyond the experimental design maximum (only 1.0% increase in 95% quantile, but a further 43% with the no suction check off). This indicates that the prior maximum on v_{max} is also sufficiently large. The effect appears to be due to a topographical constraint. In most regions the grounding line reaches a rise in bedrock elevation that temporarily slows retreat, either because the Schoof condition is not met, or because the 'no suction' effect is stronger where the ice is thin. With these high retreat rates (3 and 5 km a^{-1}), this saturation affects most regions after two centuries. For the grounding line to retreat further, the ice sheet must become thinner (by inland effects) – decreasing the model flux so the Schoof flux condition allows the grounding line to pass over bumps – or the effective basal friction coefficient at the grounding line must decrease so the Schoof flux increases. Increasing v_{max} does affect the contribution at 2100 (by 15%), because the effect is a long-term saturation. However, the effect in the ensemble at 2100 would be far smaller than in these extreme "immediate onset" simulations.

We therefore conclude the prior distributions on the retreat parameters α_{low} and v_{max} are sufficiently wide.

2.2.4 Calibration choices

We apply the ASE calibration to the whole ice sheet because we assume the parameterisation of the grounding line retreat mechanism is universal – as is standard, for example, in calibration of general circulation models with incomplete observations[39]. In other words we assume that, for a given sliding law regime, the dependence of grounding line retreat rate on effective basal friction coefficient is the same everywhere. (We do not expect the form of the sliding law itself to be universal, which is why we do not calibrate the sliding law exponent: Section 1.7). In any case we expect the dominant effect of the calibration to be on the ASE contribution, as it is the largest part of the total, particularly at 2100.

The region most likely to behave differently is the Siple Coast, which has relatively flat bedrock topography, rather than valleys as for all other major ice streams. It is difficult to say whether this would make it more responsive or less. However, the aim of using an ensemble approach, Bayesian calibration, and a conservatively large model error term is to retain a substantial degree of modelling uncertainty in the projections to account for structural uncertainties such as this. In addition, given current understanding we assess the Siple Coast has a relatively low risk of MISI triggering (Section 1.5).

Leaving the Siple Coast contribution uncalibrated would make little difference: the prior and posterior 95% quantiles are only 0.2-0.3 cm different at 2100 and 2200. We approximate the effect of relaxing the assumption of universality everywhere – i.e. calibrating only the Amundsen Sea basin results – by assuming this basin is independent from the others. We add the posterior 95% quantile for the ASE to the *prior* 95% quantile for the sum of the other basins (see Supplementary Data for values). The estimated effect is to add approximately 6 cm (22%) at 2100 and 21 cm (29%) at 2200 to the 95% quantiles.

We test the sensitivity to model error σ_m and find it has a small effect on the 95% quantile at 2200. Decreasing σ_m (being less tolerant of model error) increases the impact of calibration, concentrating the likelihood on a smaller number of ensemble members and giving narrower uncertainty estimates: lowering the Cantelli bound from 5% probability of a positive trend to 1%, which lowers σ_m from 13.0 Gt a^{-1} to 4.6 Gt a^{-1} (from 22% to 8% of the observed mass trend), decreases the 95% quantile at 2200 from 72 cm to 68 cm. Conversely increasing the bound to 10% (σ_m = 19.3 Gt a^{-1}, or 33% of the signal) raises the 95% quantile to 74 cm. We also test the effect of changing the starting dates in the scaling factor and simulated trend calculations to be two years earlier (2002 for PIG and 2004 for TG): this lowers the 95% quantile at 2200 from 72 cm to 69 cm, with 2 cm of this arising from the scaling factors and 1 cm from the simulated trend calculation.

2.3 Summary and future directions

We show the main features of the sensitivity tests for the plastic sliding law in Figure ED10 for 2100 (left) and 2200 (right). These show the large and opposing effects of the zero buttressing 'No suction' check and bedrock slope tolerance 'Schoof flux' constraint. The largest effect that does not relate to unphysical behaviour is that of 'immediate onset' of all basins.

In terms of future directions, uncertainty in the projections could be reduced if the dominant sliding law in each region were estimated and used to generate a map of the sliding law exponent m in the model. For the retreat parameterisation we chose a simple piecewise linear function, but a more sophisticated model could use bedrock slope to account for any dependence of this relationship on topography (Section 2.2.4).

We do not vary the surface mass balance in the simulations. High resolution modelling with BISICLES[35] indicates that dynamical responses are not influenced by projected accumulation changes, but do show sensitivity to the three different initial accumulation rates to which these changes are added. The influence of initial accumulation on projections should therefore be incorporated into formal uncertainty quantification in future studies. While ASE rates in this study are generally consistent with ours (e.g. 25-95% quantiles from 2020s to 2120s for the three initial states; one projection in the range 30-70% after this), their maximum rates occur in the 2140s and end of the 22nd century, later than ours (late 21st century for linear law; mid-21st century for nonlinear and plastic). The differences appear to arise from the timing of Thwaites Glacier collapse and subsequent migration of the grounding line over a threshold in bedrock topography and friction, but could also be due to differences in timing and relative contributions of grounding line retreat and thinning. Further analysis of rate as a function of grounding line position would be needed to investigate these differences.

Finally, in this study we do not vary the two enhancement factor parameters that are set during initialisation. This would generate multiple initial β' maps and initial states, i.e. adding a multiplicative factor to the ensemble size. This was not possible for this study – though the effect of uncertainty in β' on the response is

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partially sampled with the parameterisation – but future work could investigate this in more detail.

3 Supplementary Data

3.1 Supplementary Video 1

Animation of the Amundsen Sea Embayment. Summary of the projections in ten year time steps. The black contour shows the projected median grounding line position. The map shows the mean change in surface elevation, with -100 m contour shown in green. Dashed purple lines show the borders of Pine Island Glacier (PIG) and Thwaites Glaciers, which together comprise the Amundsen Sea Embayment.

3.2 Supplementary Table

Excel file of sea level projections. Columns:

```
Region Name Dist Year Prob SLE Rate
```

- Region: ALL for Antarctica, B1 to B14 for individual basins
- Name: abbreviated basin names (see full names below)
- Dist: Prior (uncalibrated) or Posterior (calibrated) distribution
- Year: 2050, 2100, 2150, 2200
- Prob: probability (%) that sea level contribution (SLE) and/or rate (Rate) is below value in following two columns (see list below)
- SLE: projected contribution in cm sea level equivalent
- Rate: projected rate of contribution in mm per year sea level equivalent

Region names (basin definitions same as in BISICLES study[35]):

```
ALL ANT: Antarctica
B1  PEN: Peninsula
B2  ASE: Amundsen Sea Embayment
B3  MBL: Marie Byrd Land
B4  SIP: Siple Coast
B5  ELL: Ellsworth
B6  PMN: Pensacola Mountains
B7  SHA: Shackleton Range
B8  DML: Dronning Maud Land
```

```
B9  END: Enderby Land
B10 MRB: MacRobertson Land
B11 PEL: Princess Elizabeth Land
B12 WIL: Wilkes Land
B13 GVL: George V Land
B14 TMN: Transantarctic Mountains
```

Probability values are given for particular quantiles, and thresholds of SLE and rate:

1. Quantiles (%): 5, 10, 25, 33, 50, 66, 75, 90, 95
2. Probabilities of not exceeding SLE thresholds (cm): 1, 2, 3, 5, 10, 20, 30, 40, 50, 100
3. Probabilities of not exceeding rate thresholds (mm a^{-1}): 0.1, 0.2, 0.3, 0.5, 1.0, 1.2, 1.3, 1.4, 1.5, 2

The second and third groups are 100% - (exceedance probability).

3.3 Supplementary Data

R script (projections.R) and data file (Ritz_et_al_Nature_projections.RData) to output annual Antarctic posterior projections. Options: modes (cm SLE), quantiles (cm SLE) or probabilities of non-exceedance (%) for any quantile/threshold and date range. These results correspond to the columns 'Prob' (probabilities of non-exceedance) and 'SLE' (modes, quantiles) for rows 'ALL ANT Post' in the Excel file, but using this script any quantile/threshold and years can be selected.

3.4 Code availability

Code for the retreat rate parameterisation and to reproduce non-map figures (Figs. 2, 3, ED1, ED5-10) and Supplementary Data is available on request: please contact Tamsin Edwards (tamsin.edwards@open.ac.uk).

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